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R. Zech, C. Kull, P. W. Kubik, H. Veit. Exposure dating of Late Glacial and pre-LGM moraines in the Cordon de Doña Rosa, Northern/Central Chile (?31° S). *Climate of the Past Discussions*, 2006, 2 (5), pp.847-878. hal-00298151

HAL Id: hal-00298151

<https://hal.science/hal-00298151>

Submitted on 25 Sep 2006

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Exposure dating of Late Glacial and pre-LGM moraines in the Cordon de Doña Rosa, Northern/Central Chile ($\sim 31^\circ$ S)

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Received: 10 August 2006 – Accepted: 5 September 2006 – Published: 25 September 2006

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Abstract

Despite the important role of the Central Andes (15–30° S) for climate reconstruction, knowledge about the Quaternary glaciation is very limited due to the scarcity of organic material for radiocarbon dating. We applied ^{10}Be surface exposure dating (SED) on 22 boulders from moraines in the Cordon de Doña Rosa, Northern/Central Chile (~31° S). The results show that several glacial advances in the southern Central Andes occurred during the Late Glacial between $\sim 14.7 \pm 1.5$ and 11.6 ± 1.2 ka BP. A much more extensive glaciation is dated to $\sim 32 \pm 3$ ka BP, predating the temperature minimum of the global LGM (Last Glacial Maximum: ~ 20 ka BP). Reviewing these results in the paleoclimatic context, we note that the Late Glacial advances coincide with (i) lower temperatures during the Younger Dryas (YD) and the Antarctic Cold Reversal (ACR), (ii) the intensification of the tropical circulation and a corresponding increase in summer precipitation and (iii) a minimum in austral summer insolation favouring reduced ablation. The absence of LGM moraines indicates that moisture advection was too limited to allow significant glacial advances at ~ 20 ka BP. The tropical circulation was much less intensive despite the maximum in austral summer insolation. Winter precipitation was apparently also insufficient, although pollen and marine studies indicate a northward shift of the westerlies at that time. The dominant pre-LGM glacial advances in Northern/Central Chile at ~ 32 ka BP required lower temperatures and increased precipitation than today. They coincide with (i) a minimum of southern high-latitude insolation suggesting an equatorward shift of the westerlies due to increased snow and ice cover, (ii) a maximum winter insolation resulting in ocean-continental temperature and pressure gradients favouring moisture advection, (iii) minimum summer insolation suggesting lower temperatures and reduced ablation and (iv) low high-latitude temperatures corroborating that they affect subtropical and tropical temperatures. More glacier-climate modelling is necessary to quantify the influence of the various forcings on the dated glacial advances.

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1 Introduction

Surface exposure dating is a new, innovative method based on in-situ production and accumulation of so-called cosmogenic nuclides (e.g. ^{10}Be , ^{26}Al and ^{36}Cl) within the first few decimetres of an exposed rock surface (Gosse and Phillips, 2001). It can be applied to determine the deposition age of boulders and moraines, respectively. Several recent studies have shown that SED is already a powerful tool for glacial and climate reconstructions (e.g. Briner et al., 2005; Owen et al., 2005; Shulmeister et al., 2005; Smith et al., 2005).

So far, very little is known about the glaciation history in the Central Andes, mainly due to the lack of organic material for radiocarbon dating (Harrison, 2004; Heine, 2004; Veit, 2006). The Central Andes play, however, an important role in Quaternary climate reconstruction, because they are situated at the transition zone between the tropical and the mid-latitude atmospheric circulation system of the PEP I transect (Pole-Equator-Pole American Transect: <http://www.pages.unibe.ch>, Markgraf et al., 2000). Changes in the intensity or latitudinal shifts of the circulation systems should be recorded in suitable archives and provide valuable information about forcings and mechanisms of the climate system.

We have previously published results from the nearby Encierro Valley ($\sim 29^\circ\text{S}$, 11 samples), which indicated that a prominent glacial advance occurred during the Late Glacial ($\sim 14.0 \pm 1.4\text{ ka BP}$) (Zech et al., 2006). Several recessional moraines were deposited before deglaciation at the Pleistocene-Holocene transition ($\sim 11.6 \pm 1.2\text{ ka BP}$). An earlier maximum glacial advance ($> 24.1 \pm 2.4\text{ ka BP}$) had tentatively been postulated based on a single boulder. Here, we now present 22 additional ^{10}Be exposure ages from moraines in the Cordon de Doña Rosa, Northern/Central Chile ($\sim 31^\circ\text{S}$, Fig. 1). The aim of the present study is to evaluate the previous results from the Encierro Valley. Specifically, we tried to (i) specify the succession of the Late Glacial advances and (ii) to determine the timing of the earlier, more extensive glacial advances. In addition, (iii) the paleoclimatic context has been reviewed in more detail in

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order to identify possible forcings on the glaciation in the southern Central Andes.

2 Climate and paleoclimate of Northern/Central Chile

Our research area is situated south of the “Arid Diagonal” (Fig. 1). This is the transition zone between areas in the Central Andes that are influenced by the westerlies (to the south) and areas that are influenced by the tropical circulation system (to the north). The westerlies provide moisture from the Pacific. Their seasonal northward shift results in an austral winter maximum. A steep precipitation gradient characterizes the Andean rainfall at 4000 m altitude: annual precipitation exceeds 4000 mm at $\sim 40^\circ$ S, but rapidly decreases to ~ 2000 mm at $\sim 35^\circ$ S. Only 400 mm/a are available at $\sim 30^\circ$ S, and rainfall further decreases to less than 100 mm/a at 26° S (Vuille and Ammann, 1997). In the centre of the Arid Diagonal aridity becomes so severe that no glaciers exist despite altitudes >6000 m a.s.l. (above sea level): that is between 18° S (Vulcan Sajama) and 27° S (Cerro Tres Cruces) (Ammann et al., 2001). The fact that glaciers are limited by precipitation rather than temperature has been described as “thermal readiness” (Messerli, 1973). Eventually, north of the Arid Diagonal, the tropical circulation system provides moisture from the Atlantic and the Amazon Basin (Zhou and Lau, 1998: “South American Summer Monsoon” (SASM); Garreaud et al., 2003; Vuille and Keimig, 2004).

Due to the steep humidity gradient, our research area is a key site for the reconstruction of changes in the intensity and/or shifts in the atmospheric circulation patterns. Nevertheless, knowledge about Late Quaternary glacial and climate changes in the southern Central Andes is still very limited (Garleff and Stingl, 1991, 1998; Veit, 1995, 2006; Espizua, 2004; Harrison, 2004). Large moraines have been mapped both north and south of the Arid Diagonal (Jenny and Kammer, 1996; Ammann et al., 2001). They document Late Pleistocene phases of substantially increased precipitation, but age control is almost absent: North of the Arid Diagonal, they have been correlated with the Late Glacial lake transgression phases, which are well-documented on the

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Altiplano (“Tauca”: ~16–14 ka BP; “Coipasa”: ~13–11 ka BP) (Clapperton et al., 1997; Clayton and Clapperton, 1997; Geyh et al., 1999; Sylvestre et al., 1999; Grosjean et al., 2001; Placzek et al., 2006). Evidence for a Late Glacial humid phase disappears south of 25° S, both concerning lacustrine and glacial deposits (Geyh et al., 1999; Ammann et al., 2001). Amman et al. (2001) concluded that the maximum glacial advances north and south of the Arid Diagonal might not have occurred synchronously. The same conclusion has been deduced from a glacier-climate model that indicated a temperature depression of ~5.5°C for the prominent moraine (M-II) in the Encierro Valley (Kull et al., 2002). This temperature depression is much stronger than the ones calculated for the Western Cordillera at ~18° and 23° S (2–3°C), where the modelling results indicated very wet, relatively warm “Late Glacial” climate conditions (Kull, 1999; Kull and Grosjean, 2000). No age controls, however, have been available so far to test whether the prominent moraines south of the Arid Diagonal correlate with the LGM advances dated in the Chilean Lake District (Lowell et al., 1995; Denton et al., 1999).

As our previous dating efforts in the Encierro Valley revealed a Late Glacial age for the prominent M-II moraine (Zech et al., 2006: ~14.0±1.4 ka BP) but did not allow to determine the timing of the earlier, more extensive glaciation with certainty (>24 ka BP), we decided to sample moraines for ¹⁰Be SED in the nearby Cordon de Doña Rosa.

3 Cordon de Doña Rosa/sampling

The Cordon de Doña Rosa (~30°42' S, 70°26' W, Fig. 2) lies ~200 km south of the Encierro Valley and ~50 km east of the town Monte Patria. The summits reach altitudes of ~4500 m a.s.l., but no glaciers exist there today. Large prominent lateral moraines extend down to ~3000 m a.s.l., where they just reach the east-west stretching main valley of the Rio Los Molles (DR5, DR6 and DR7, Fig. 3 Photo 1). Farther downstream, the U-shaped-valley indicates that a former, extensive glaciation reached as deep as <2000 m a.s.l. (DR8, DR9 and DR10, Fig. 3 Photo 2). A complex pattern of

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glacial deposits can be found up-stream (>3300 m a.s.l., DR1-4, Fig. 3 Photos 3 and 4). These recessional moraines are mostly not well-defined crest-like features, but rather flat and smooth.

We sampled (i) the prominent lateral moraines (DR5, DR6 and DR7), (ii) boulders in the main valley (DR8, DR9 and DR10) and (iii) the recessional moraines (DR1-DR4) for ^{10}Be SED (see Fig. 2 for the exact sampling location). Approximately 0.5 kg rock surface material was chipped from the top of the granite boulders with hammer and chisel. We followed standard sampling strategies, i.e. preferring large un-eroded boulders in a stable position. All samples were documented by photograph (see supplementary material: <http://www.clim-past-discuss.net/2/847/2006/cpd-2-847-2006-supplement.pdf>), GPS (latitude, longitude, altitude) and a field description (including topographic shielding). The laboratory work followed standard procedures (e.g. Ivy-Ochs, 1996), i.e. in brief: separation of the quartz, dissolution in HF after a ^9Be carrier has been added, purification of beryllium using anion and cation exchange chromatography, precipitation and oxidation of beryllium (A detailed laboratory procedure is provided as supplementary material in Zech et al., 2006b¹). The AMS measurements were conducted by P. Kubik at the ETH/PSI tandem facility in Zurich.

4 Results and discussion

Exposure ages were calculated using the scaling system of Desilets and Zreda (2003) and a reference production rate of $5.25 \text{ atoms a}^{-1} \text{ g}^{-1} \text{ SiO}_2$ for neutron spallation. This value has been derived from recalculation of cosmogenic nuclide data of the Kofels landslide in Austria (Kubik et al., 1998; Kubik and Ivy-Ochs, 2004, and three unpublished data from U. Abramowski, personal communication). Fast and slow muon

¹Zech, R., Kull, C., and Veit, H.: Potentials and limitations of surface exposure dating for glacial and climate reconstruction in the Central Andes, Quaternary Geochronology, submitted, 2006b.

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production rates at sea level and high latitude are 0.09 and 0.11 atoms $\text{a}^{-1} \text{g}^{-1}$, respectively (Heisinger et al., 2002a, b). All exposure ages are corrected for changes of the magnetic field intensity (data from McElhinny and Senanayake, 1982, for the time <10 ka BP; and from Guyodo and Valet, 1996, for the time >10 ka BP). The dipole wobble is accounted for during the last 10 ka (data according to Ohno and Hamano, 1992). Before that time, a geocentric axial dipole can be assumed without inducing major errors. A Microsoft[®] Excel calculation scheme is provided as supplementary material in Zech et al. (2006)¹. All sample data and calculated exposure ages are given in Table 1 and illustrated in their stratigraphical context in Fig. 2.

4.1 Sample-specific effects and uncertainties

Exposure ages of boulders do not necessarily reflect the deposition age of the respective moraine. Too old boulders can be explained with “inheritance”, i.e. pre-exposure. In general, inheritance is assumed to affect less than 3% of the samples (Shanahan and Zreda, 2000; Putkonen and Swanson, 2003). On the other side, too young samples can be the result of rock surface erosion or post-depositional landform surface instability: A recent chipping event of 10 cm, for instance, leads to an underestimation of the deposition age of ~10%. Concerning landform surface instability, formerly shielded rock surfaces can be exposed by a wide range of geomorphological processes, like denudation of the surrounding soil matrix (Hallet and Putkonen, 1994; Putkonen and Swanson, 2003), upheaval or rotation of boulders due to cryoturbational processes, and long-lasting ice-decay in case of ice-cored moraines (Zech et al., 2005).

One can try to identify too old or too young samples based on the overall stratigraphic and chronostratigraphic situation. For this purpose, we plot all exposure ages in the order of the anticipated relative chronology, i.e. from younger to older from left to right (Fig. 4). The data scatter and the inconsistencies are due to the aforementioned sample-specific effects and the AMS measurement uncertainties (error bars). Note that negligible sample-specific effects were omitted for simplicity. These are topo-

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graphic shielding (<1%, calculated according to Dunne et al., 1999), sample thickness (negligible for chips <4 cm, Masarik and Reedy, 1995), and vegetation and snow cover (the vegetation cover is extremely sparse and preferential sampling of large boulders minimizes snow cover effects).

5 4.1.1 Late Glacial moraines

Two boulders from the stratigraphically youngest sampled moraine have exposure ages of 11.6 and 12.8 ka BP (DR11 and DR13, respectively. Note that the AMS errors are omitted in the text for reasons of readability). The two ages are indistinguishable within the AMS measurement uncertainty. A single boulder has been dated from the terminal moraine DR2. Its exposure age of 17.5 ka BP is rejected, because it is inconsistent with the six younger exposure ages from the moraines DR3 and DR4 down-valley (→ assumed inheritance). The terminal moraines DR3 and DR4 are dated with three exposure ages each to 11.7 (DR31), 13.2 (DR33) and 14.7 ka BP (DR32), and to 11.7 (DR41), 12.1 (DR43) and 12.5 ka BP (DR42).

15 In general, the application of the “oldest age model” can be recommended (Zech et al., 2005). It suggests that the oldest exposure age from a deposit is the best available estimate for the deposition age, unless there are stratigraphical inconsistencies. Unfortunately, this is the case for the Late Glacial moraines in the Cordon de Doña Rosa. Here, DR4 would be younger than DR3, and both would be younger than DR2. We therefore conservatively conclude that due to the sample-specific effects (especially potential inheritance vs. erosion/landform surface instability), it seems to be impossible to determine the exact deposition age of the individual Late Glacial moraines. Glacial advances occurred between ~11.6 and 14.7 ka BP. Note that rejecting DR21 (17.5 ka BP) does not alter the general paleoclimatic interpretation. Eventually, Holocene warming (at ~11.6 ka BP) initiated deglaciation, melting of buried glacier ice and stabilization of the landform surfaces.

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4.1.2 Pre-LGM moraines

The inner prominent latero-frontal moraine was sampled close to a debris fan (DR5) and on a crest-like section (DR6). Three of four boulders date the corresponding glacial advance to ~32 ka BP (DR62: 31.1, DR52: 32.1 and DR61: 34.4 ka BP). The outlier DR51 (98.1 ka BP) likely contains inherited ¹⁰Be. The outer prominent moraine (DR7) has apparently been affected more intensively by erosion. This is indicated by the smoother morphology and signs of weathering on many boulders. In fact, none of the three sampled boulders on DR7 reaches an exposure age older than the stratigraphically younger inner lateral moraine. Closest is the stable, big boulder DR71 (29.8 ka BP). Boulder DR73 (25.1 ka BP) is obviously eroded, and boulder DR72 (15.6 ka BP, Photo DR72) is in an unstable position and might have toppled. The applicability of SED seems to have reached its limits (set by erosion and landform surface instability).

4.1.3 Maximum glaciation

Exposure ages from boulders on moraine remnants of the most extensive glaciation were obtained from three sites in the U-shaped main valley (DR8, DR9 and DR10). The ages range from 16.8 (DR101) to 43.1 ka BP (DR82). As evident from Fig. 4, only two ages are not inconsistent regarding the stratigraphical context. In general, all boulders show signs of weathering and are not ideally suited for SED. We conclude that it is not possible to determine the age of the most extensive glaciation. It must have occurred before 32 ka BP, probably much earlier, because of the intensive erosion of the moraine remnants.

4.2 Systematic uncertainties – absolute glacial chronology

One should be aware of the current limitations of the dating method when discussing the exposure ages in the paleoclimatic context. Systematic uncertainties, i.e. those affecting all samples in the research area in the same way, are mainly due to uncer-

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ainties concerning the reference production rate (at sea level and high latitude, SLHL, provided by calibration studies) and the scaling system. Scaling is necessary to calculate the local, effective production rate from the reference production rate and thus to account for the altitude and latitude dependence of the cosmic radiation. There is

neither consensus yet concerning the exact value of the reference production rate, nor concerning scaling. A major international effort has started recently with the CRONUS-project aiming at the development of an internationally accepted protocol that will ultimately allow a precision of SED better than 5% (CRONUS-Earth and CRONUS-EU projects: <http://www.physics.purdue.edu/cronus>, and <http://www.cronus-eu.net>, respectively). In the meantime, it is appropriate to illustrate the sensitivity of the exposure ages on various calculations schemes and assumptions. This is accomplished in Fig. 5, exemplarily for the samples DR11 and DR71.

The first columns represent the exposure ages as presented above, i.e. using the scaling system of Desilets and Zreda (2003), a reference production rate of $5.25 \text{ atoms a}^{-1} \text{ g}^{-1} \text{ SiO}_2$ for neutron spallation and geomagnetic corrections. The 10% error bars illustrate the total systematic uncertainties as suggested by Gosse and Phillips (2001). The scaling system of Dunai (2001) (second columns, re-calculated total reference production rate: $5.51 \text{ atoms a}^{-1} \text{ g}^{-1} \text{ SiO}_2$, muon contribution 3.6%) yields slightly higher exposure ages. Our results are in very good agreement with the calculation scheme provided by Pigati and Lifton (2004) (third column, using a re-calculated total reference production rate: $5.42 \text{ atoms a}^{-1} \text{ g}^{-1} \text{ SiO}_2$). As illustrated by the fifth columns, calculations using the scaling system of Stone (2000) result in much older exposure ages. This is partly due to the fact that geomagnetic corrections are not included in that protocol. In order to enable direct comparison we therefore also did our own calculations (using the scaling system of Desilet and Zreda) excluding the geomagnetic corrections (fourth columns). This reveals that most of the offsets are due to other reasons, like the different cosmic ray dataset and the parameterisation of the neutron flux. The last columns show that even a high hypothetical constant uplift rate of 5 mm/a would not lead to dramatic underestimation of the exposure ages and can therefore be

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neglected.

We emphasise that arguments in favour of the scaling system of Desilet and Zreda (2003) are (i) that it is based on the best-characterized cosmic-ray dataset, (ii) it takes into account the altitude dependence of the energy spectrum of the cosmic radiation and (iii) it explicitly includes corrections for changes in the geomagnetic field. Nevertheless, our results will be subject to minor changes in the future. An improved understanding of the underlying physics will reduce the systematic uncertainties, and local calibration studies will be indispensable to address long-term atmospheric pressure anomalies and anomalies in the geomagnetic field. In the following, we will quote our ages with the estimated 10% systematic error according to Gosse and Phillips (2001).

The results of the present study corroborate and supplement the glacial chronology that we have previously derived from SED in Northern/Central Chile (Zech et al., 2006):

- Glaciers were much more extensive than today during the Late Glacial between $\sim 14.7 \pm 1.5$ and 11.6 ± 1.2 ka BP. Geomorphological evidence shows that several moraine stages were deposited by fluctuating glaciers. As mentioned above, the sample-specific uncertainties impede the refinement of the Late Glacial chronology.
- A large, well-defined latero-frontal moraine is dated to $\sim 32 \pm 3$ ka BP. The timing of earlier, even more extensive glaciations cannot be determined any more with SED due to erosion.
- We find no evidence for prominent glacial advances in Northern/Central Chile during the global LGM. Of course, this does not rule out that glaciers advanced at that time, but they were limited compared to the other, datable stages.

4.3 Paleoclimatic context

In the following, we discuss the glacial chronology of Northern/Central Chile in context with other paleoclimate records: first with a focus on the tropical circulation, then on

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the westerlies. Finally, we consider possible direct insolation forcings and the role of high-latitude temperature. Thus, we try to identify possible forcings on the glaciation and to contribute to the overall understanding of the Late Pleistocene climate changes in the Central Andes.

5 4.3.1 Late Glacial advances and the increase in summer precipitation

For the 14 ka BP event in the Encierro Valley detailed glacier-climate modelling showed that not only a temperature depression of $\sim 5.5^{\circ}\text{C}$, but also an increase of annual precipitation by ~ 580 mm (today 400 mm) was necessary (Kull et al., 2002). Interestingly the model could not determine the seasonality of precipitation, although today predominately winter precipitation prevails. For the reconstruction of the atmospheric circulation, it is of particular importance to determine the moisture source.

Whereas there is not much information about the Late Glacial humidity conditions south of our research area, plenty of evidence proves significantly wetter Late Glacial/Early Holocene conditions both on the Bolivian Altiplano and as far south as 28°S (Geyh et al., 1999; Baker et al., 2001a, b; Grosjean et al., 2001; Placzek et al., 2006). Reservoir effects complicate the exact timing of the transgression phases, but the most recent and extensive study applied U/Th dating in combination with radiocarbon, showing that high shorelines on the Bolivian Altiplano were formed between 16–14 (“Tauca”) and between 13–11 ka BP (“Coipasa”) (Placzek et al., 2006). Glacier advances on the Altiplano have been correlated with that humid phase (Clapperton et al., 1997: Choqueyapu II glacial advances at 14–13 ka BP and 12–10 ka BP, uncalibrated 14C ages). We have already previously argued that the Late Glacial advances in Northern/Central Chile at $\sim 30^{\circ}\text{S}$ may have been caused by the increase in tropical moisture as well (Zech et al., 2006a). This interpretation is in good agreement with pollen analyses from rodent middens in Chile at $\sim 25^{\circ}\text{S}$, which indicate increased summer precipitation between 17 and 11 ka BP (Maldonado et al., 2005). Moreover, speleothem data from SE-Brazil can be interpreted to document the intensification of the South American Summer Monsoon (SASM) during the Late Glacial (Wang et al.,

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2006; Cruz et al., 2005, 2006, Fig. 6a). As illustrated in Fig. 6, relatively negative $\delta^{18}\text{O}$ values in SE-Brazil correlate with more positive values in the northern hemispheric speleothems (Wang et al., 2001; Yuan et al., 2004). This documents the anti-phasing of the monsoonal precipitation, which is controlled by the hemispheric asymmetry in insolation forcing on orbital timescales and which likely also reflects millennial-scale latitudinal shift of the ITCZ due to high-latitude boundary conditions. The mechanism to transport the tropical/subtropical humidity from the South American lowlands onto the highlands of the Central Andes has been identified in both climate modelling and observational studies: upper-tropospheric easterly anomalies, which result from an intensification and a southward shift of the Bolivian High (Garreaud et al., 2003; Vuille and Keimig, 2004).

4.3.2 Northward shift of the westerlies during the LGM and/or the pre-LGM?

As evident from the speleothem data in SE-Brazil (Cruz et al., 2006: see Fig. 6) and the shoreline studies on the Altiplano (Placzek et al., 2006) the tropical circulation system was less intensive during the LGM than during the Late Glacial. This is reflected by the absence of corresponding moraines in our research area. It was obviously too dry to allow dominant glacial advances despite the global temperature minimum. Apparently, the westerlies did not provide sufficient moisture either. Several studies have postulated a northward shift of the westerlies and increased winter precipitation in Northern/Central Chile during the LGM (e.g. Heusser, 1989; Stuut and Lamy, 2004; Maldonado et al., 2005). There are, however, still unresolved inconsistencies and controversies concerning these findings (Markgraf, 1989; Grosjean et al., 2003; Stuut et al., 2006). Climate modelling can not yet unambiguously contribute to this issue (e.g. Valdes, 2000; Wyrwoll et al., 2000). The timing and the extent of the latitudinal shifts of the westerlies clearly deserve further research.

In any case, much more precipitation than today and during the LGM was required in combination with low temperatures to trigger the massive pre-LGM advances at

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~32 ka BP in Northern/Central Chile (preliminary results from glacier-climate modelling, C. Kull, 2006, personal communication). Significant contribution of tropical moisture can be excluded for that time (Cruz et al., 2006; Placzek et al., 2006), so that a northward shift and/or intensification of the westerlies can be inferred. There is glacial evidence for synchronous advances along the Central Andes as far south as ~40° S: (1) At 36° S $^{40}\text{Ar}/^{39}\text{Ar}$ dating of glaciated and unglaciated lava flows shows that pre-LGM ice caps existed and retreat started already between 25.6 and 23.3 ka BP (Singer, 2000). (2) Preliminary exposure ages from moraines in the Valle Rucachoroi (~39° S, Argentina) indicate that the maximum glaciation occurred at ~32 ka BP (ongoing own research). (3) Extensive radiocarbon dating allowed Denton et al. (1999) to demonstrate that several glacier advances occurred between ~29 and 15 ^{14}C ka BP in the Chilean Lake District (41–43° S). Whereas the northern parts of the Chilean Lake District experienced maximum glaciation at ~22 ^{14}C ka BP, the glaciers in the southern parts were most extensive at ~15 ^{14}C ka BP. From the overall pattern of glaciation we tentatively conclude that the westerlies, which have their focus at ~50° S today, may have been shifted northward during the LGM with a focus at ~45° S (e.g. Heusser, 1989, 2003; Singer, 2000; Stuut and Lamy, 2004), but they seem to have been much more intensive and/or shifted north at ~32 ka BP. This is consistent with geochemical and stable isotope analyses from the Laguna Tagua Tagua in Central Chile showing wet conditions from ~40 to 20 ka BP (Valero-Garcés et al., 2005).

An explanation for the suggested shifts of the westerlies may partially be found in the Antarctic sea ice extent, pushing the westerlies equatorwards. Mosola and Anderson (2006) and Anderson et al. (2002) dated marine sediments immediately overlying glacial till and concluded that at least parts of both the East and West Antarctic ice sheets reached their maximum extent before the LGM. We speculate that the minimum of the austral summer insolation at ~30 ka BP might have been the respective orbital forcing (Fig. 6b: low insolation → greater sea ice extent → northward shift of the westerlies → glacial advances in Northern/Central Chile).

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4.3.3 The role of direct insolation and temperature

Apart from precipitation, temperature plays a major role in controlling glaciation (Kaser, 2001; Kull et al., 2002, 2003). The lack of high-resolution, long terrestrial temperature records from low latitudes limits this discussion to (i) the possible role of direct insolation forcing and (ii) the influence of the high-latitudes, where ice-cores provide suitable temperature proxies.

(i) As illustrated in Fig. 6 the austral summer (December) insolation at 30° S was at a minimum at ~11.6–14.7 ka BP and ~32 ka BP, the time of the glacial advances in Northern/Central Chile. We suggest that low summer insolation lead to reduced surface temperatures and reduced ablation favoring glacier growth. On the other hand, austral winter insolation at these times was quite high (see Fig. 6b). We argue that winter temperatures are not a limiting factor for glacier growth in the Central Andes. Quite the contrary, increased winter insolation changes the ocean-continent temperature and pressure gradient, thus favoring moisture advection and accumulation of snow and ice. Similar considerations are quite common for the summer monsoon regimes (Clemens et al., 1991; Gasse, 2000; Cruz et al., 2005). A recent modeling study corroborates the importance of the precessional forcing and resulting ocean-continent pressure gradients for low-latitude precipitation responses (Clement et al., 2004).

(ii) According to the same model of Clement et al. (2004) the low-latitude temperatures react very sensitively to high-latitude glacial forcing (polar temperatures). As evident from the temperature proxies in Greenland and Antarctica (Fig. 6c), low temperatures prevailed not only at ~20 ka BP, but already since ~32 ka BP, thus allowing for the observed pre-LGM glacial advances in the southern Central Andes. Concerning the Late Glacial advances they coincide with reduced temperatures at high latitudes as well (see Fig. 6c): In Greenland a sudden return to glacial conditions is well known as the Younger Dryas (YD). In Antarctica a sudden drop in temperatures interrupts the general Late Glacial warming trend.

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This event slightly predates its northern-hemispheric counterpart and is known as the Antarctic Cold Reversal (ACR). Much controversy exists over the question whether climate and/or glacial events in South America were synchronous with the YD or the ACR (reviewed in: Harrison, 2004; Heine, 2004; Mark et al., 2004). As we have shown above, the systematic uncertainties currently make it impossible to solve that debate by applying SED. Probably both high-latitude cold events had an effect on low and mid-latitude temperatures and corresponding glacial advances.

5 Conclusions

Surface exposure dating of moraines in the Cordon de Doña Rosa, Northern/Central Chile (30° S), corroborates the results from our previous study in the Encierro Valley (Zech et al., 2006a): between 14.7 ± 1.5 and 11.6 ± 1.2 ka BP several glacial advances occurred, probably documenting rapid Late Glacial temperature changes, as well as very wet conditions at that time. Lake sediment studies on the Altiplano and speleothem data from SE Brazil show a substantial intensification of the tropical circulation system (Cruz et al., 2006; Placzek et al., 2006). Hitherto the influence of the tropical circulation on the southern Central Andes might have been underestimated, but the intensification and southward shift of the Bolivian High and the corresponding upper-tropospheric easterlies provide a possible mechanism to transport the moisture from the South American lowlands onto the highlands (Garreaud et al., 2003; Vuille and Keimig, 2004). The role of the westerlies during the Late Glacial requires further investigations. More dating efforts would be necessary to determine the detailed succession of Late Glacial events, because pre-exposure and landform surface instability complicate the interpretation of the exposure ages.

Although there is some paleoclimatic evidence for a northward shift of the westerlies during the LGM, no corresponding moraines could be dated. Apparently, it was too dry to allow significant glacial advances at that time. Instead, an older, extensive glaciation

is dated to ~32 ka BP. Increased winter precipitation can be inferred for this time. An equatorward shift of the westerlies might have been caused by the maximum extent of at least parts of the Antarctic Ice Sheets and coincides with minimum solar insolation at high southern latitudes.

5 Concerning the possible role of direct insolation forcing, we note that austral summer insolation was low at ~32 ka BP and 14.7–11.6 ka BP. Reduced summer temperatures and reduced ablation probably favored glacier growth. On the other hand, austral winter insolation was high during these time periods. Winter temperatures, however, were probably never a limiting factor. Instead, changes in the ocean-continental temperature
10 and pressure gradient and thus increased moisture advection could be inferred. High-latitude glacial forcing (low temperatures) probably also played an important role for the glaciation in Northern/Central Chile, both at ~32 ka BP and during the Younger Dryas and the Antarctic Cold Reversal, respectively.

Ongoing work applying SED aims at the determination of the spatial extent of the
15 dated glacial advances. In combination with the glacier-climate-model we intend to better quantify the role of temperature and precipitation. The model will have to consider seasonality in a more sophisticated way, so that the paleoclimatic interpretations can be further refined. Last but not least, we highlight the need for improvements in the physics of the cosmogenic radiation with regard to altitude and latitude scaling, as
20 well as the need for calibration sites in the Central Andes in order to reduce the current systematic uncertainties of SED.

Appendix A

Photographs from all sampled boulders are provided as supplementary material (<http://www.clim-past-discuss.net/2/847/2006/cpd-2-847-2006-supplement.pdf>).

25 *Acknowledgements.* We would like to thank the Swiss National Science Foundation for funding (SNF project 21-067937.02/1), and M. Vuille and N. Lifton for helpful discussions. R. Zech gratefully acknowledges the support by the “Studienstiftung des deutschen Volkes e.V.”.

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Table 1. Sample data and exposure ages calculated according to Desilets and Zreda (2003).

Sample	Latitude °S	Longitude °W	Altitude [m a.s.l.]	¹⁰ Be [10 ⁵ at/g SiO ₂]	AMS SD ^a [%]	Exposure age [ka BP] ^b
Late Glacial moraines :						
DR11	30.68	70.37	3806	7.657	3.8	11.6±0.5
DR13	30.68	70.37	3762	8.242	3.9	12.8±0.5
DR21	30.68	70.37	3734	11.283	3.9	17.5±0.6
DR31	30.68	70.36	3686	7.170	4.0	11.7±0.5
DR32	30.68	70.36	3683	9.057	3.7	14.7±0.5
DR33	30.68	70.36	3683	8.061	3.1	13.2±0.4
DR41	30.69	70.36	3614	6.895	5.3	11.7±0.6
DR42	30.69	70.36	3603	7.279	4.2	12.5±0.5
DR43	30.69	70.36	3591	7.019	3.9	12.1±0.5
Pre-LGM moraines:						
DR51	30.71	70.36	3383	58.263	3.0	98.1±2.5
DR52	30.72	70.36	3376	17.993	3.0	32.1±0.8
DR61	30.72	70.36	3316	18.806	3.6	34.4±1.0
DR62	30.72	70.36	3317	16.685	3.3	31.1±0.9
DR71	30.72	70.37	3293	15.669	5.6	29.8±1.4
DR72	30.72	70.37	3294	7.554	4.2	15.6±0.6
DR73	30.73	70.37	3300	12.855	4.1	25.1±0.9
Maximum glaciation:						
DR81	30.75	70.44	2536	10.043	4.3	31.7±1.2
DR82	30.75	70.44	2537	14.542	4.3	43.1±1.6
DR91	30.73	70.51	2058	5.920	4.4	27.2±1.0
DR92	30.73	70.51	2059	5.781	3.3	26.6±0.7
DR101	30.73	70.55	1772	2.791	10.5	16.8±1.6
DR102	30.73	70.55	1773	4.177	4.4	24.2±0.9

^a AMS SD=1 σ standard deviation of the AMS measurement^b including the propagated 1 σ standard deviation of the AMS measurement

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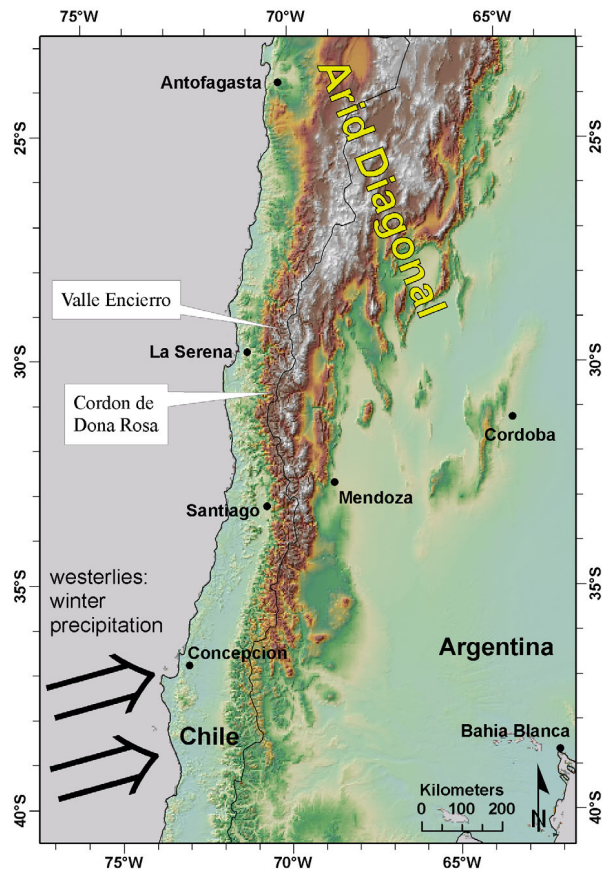


Fig. 1. Location of the research area Cordon de Doña Rosa, as well as our previous research area Valle Encierro. The westerlies provide moisture to the Central Andes south of the Arid Diagonal.

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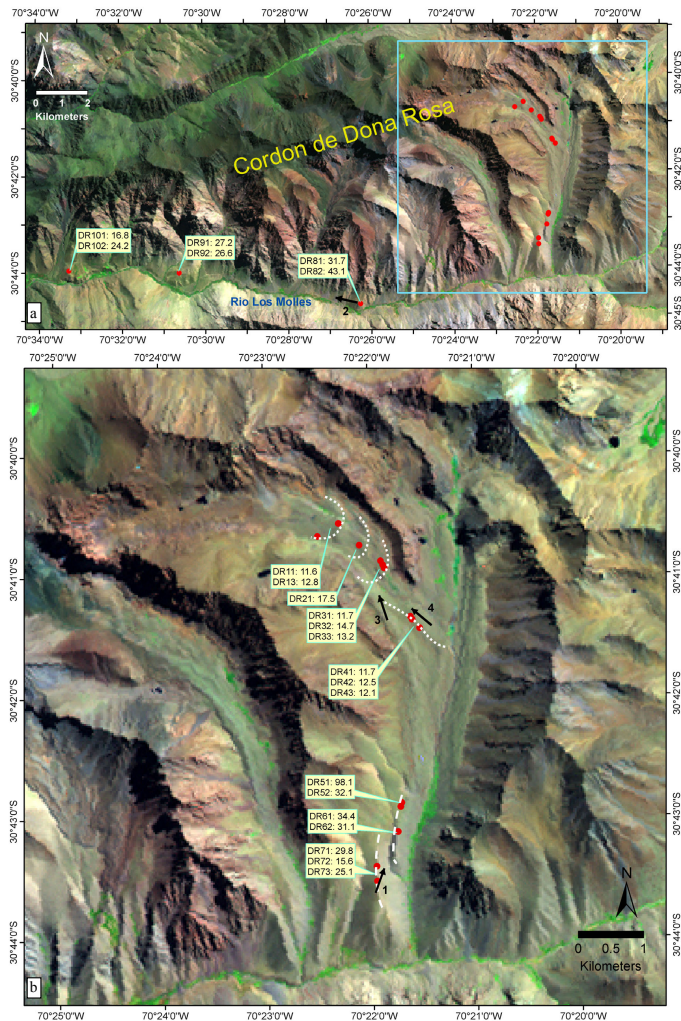


Fig. 2.

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Fig. 2. Geomorphological setting of the sampling locations (red dots) and exposure ages in ka BP. **(a)** Landsat image (RGB: 742) of the Cordon de Doña Rosa with its north-south tributary valleys disemboguing into the east-west-stretching main valley of the Rio Los Molles. **(b)** Close-up view of the prominent latero-frontal moraines and the recessional moraines. The dotted lines indicate terminal moraines, the dashed lines latero-frontal ones. The arrows indicate the standpoints and views of the photographs 1 to 4. Nomenclature DR_xy: DR = Dona Rosa, y = sampling location/moraine, x = sample number.

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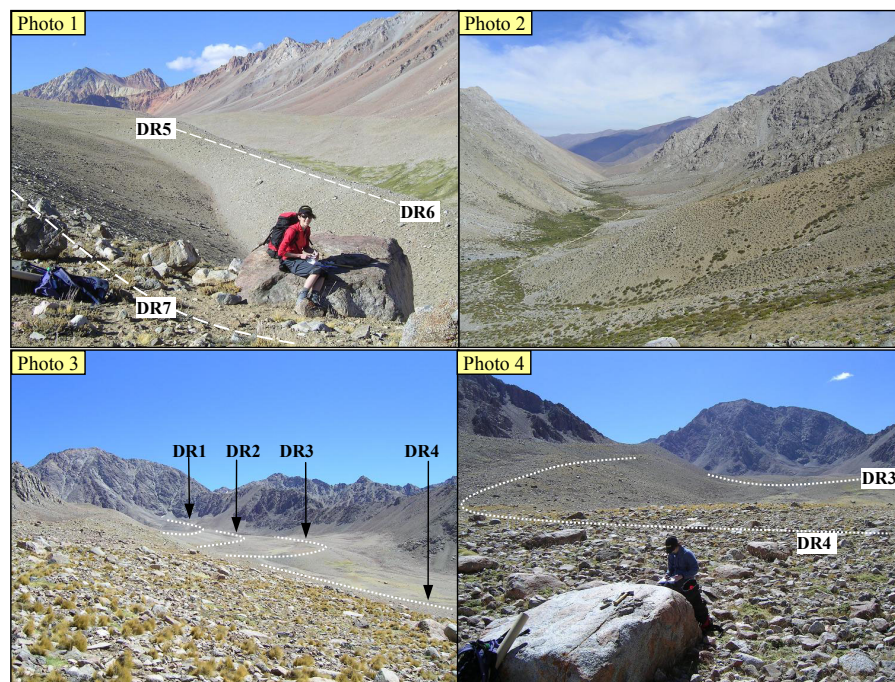


Fig. 3. Photo 1: View from the prominent latero-frontal moraine DR7 in northern direction onto the prominent lateral moraine DR6 and 5. Photo 2: View from sampling location DR8 down-valley in western direction. Note the U-shape of the valley. Photo 3: View in northern direction onto the recessional terminal moraines DR1 to 4. Photo 4: View from DR4 in north-western direction towards DR3.

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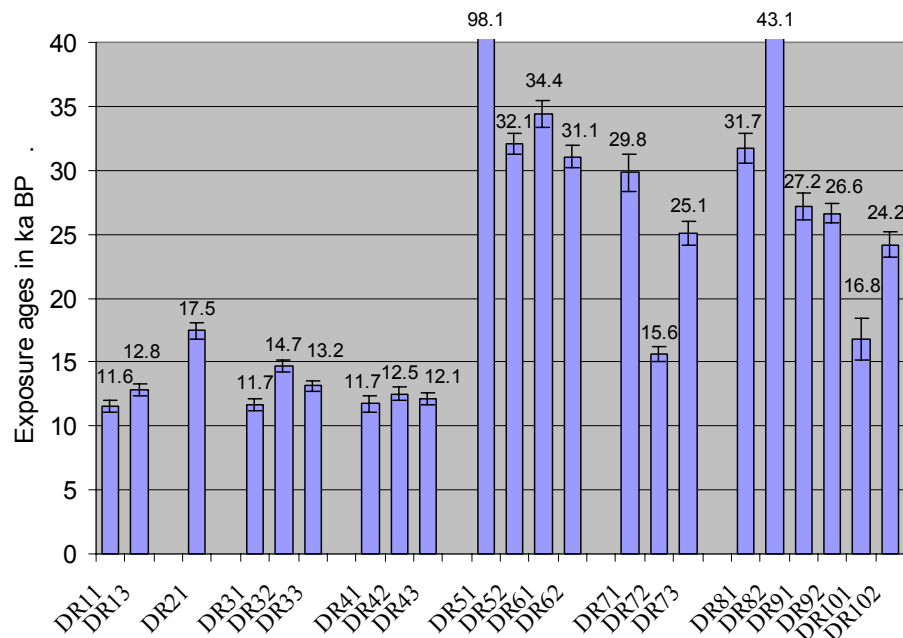


Fig. 4. Exposure age plot of all samples in the anticipated chronostratigraphic order, i.e. from younger to older from left to right. The error bars indicate the propagated 1σ AMS measurement uncertainties.

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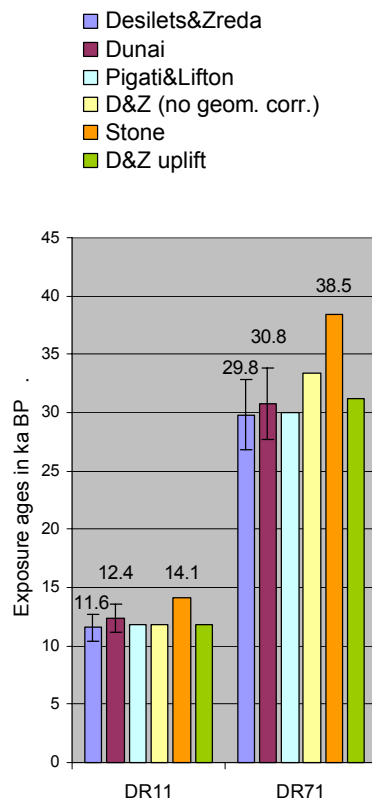


Fig. 5. Systematic uncertainties of the exposure ages. From left to right: calculation according to (1) Desilets and Zreda (2003), (2) Dunai (2001), (3) Pigati and Lifton (2004), (4) Desilets and Zreda (2003), excluding geomagnetic corrections, (5) Stone (2000) and (6) Desilets and Zreda (2003), assuming an uplift rate of 5 mm/a (including geomagnetic corrections). The 10% error bars illustrates the total systematic uncertainties as suggested by Gosse and Phillips (2001).

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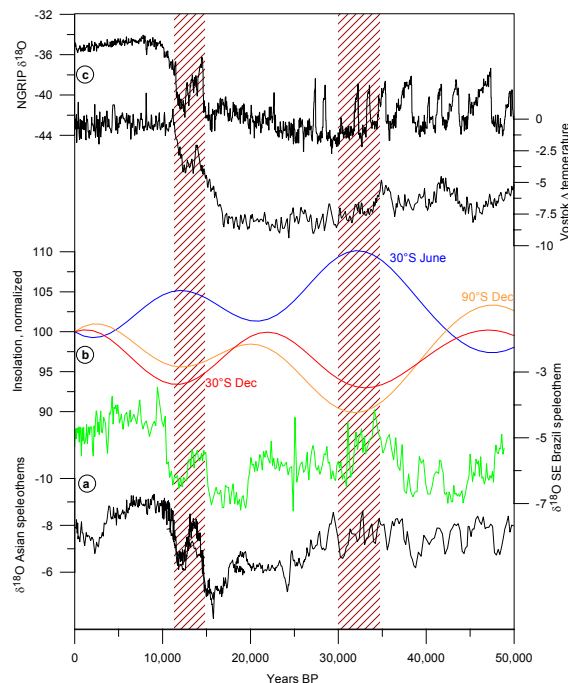


Fig. 6. Paleoclimatic context of the glacial advances in Northern/Central Chile (14.7–11.6 and ~32 ka BP, illustrated as brown shaded bars). **(a)** Speleothem $\delta^{18}\text{O}$ data: more negative values in the SE-Brazil speleothem indicate an intensification of the SASM (Cruz et al., 2006). The opposite signal in the Asian speleothem records from Hulu cave (Wang et al., 2001) and Dongge cave (Yuan et al., 2004) (reversed scale) illustrates the inter-hemispheric anti-phasing of monsoonal precipitation. **(b)** Normalized insolation changes for austral summer (30° S December and 90° S December) and austral winter (30° S June) (Berger and Loutre, 1991). **(c)** NGRIP $\delta^{18}\text{O}$ record as temperature proxy for Greenland (NGRIP members, 2004, more negative values → colder temperatures), and temperature deviation from present conditions in Antarctica derived from the Vostok deuterium record (Petit et al., 1999).

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